

Inter-decadal variation of the Tropical Atlantic – Korea (TA-K) teleconnection pattern during boreal summer season

Yoo-Geun Ham¹, YeonJi Hwang¹, Young-Kwon Lim², and Minho Kwon³

¹Faculty of Earth Systems and Environmental Sciences, Chonnam National University, Gwangju, South Korea

²Goddard Earth Sciences Technology and Research/I. M. Systems Group, Global Modeling and Assimilation Office, NASA

³Ocean Circulation and Climate Research Center, Korea Institute of Ocean Science and Technology, Ansan, Korea

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*Corresponding author: Yoo-Geun Ham, Faculty of Earth Systems and Environmental Sciences, Chonnam National University, 77 Yongbong-ro, Buk-gu, Gwangju 500-757, Republic of Korea (ygham@jnu.ac.kr)

29

Abstract

30 The inter-decadal variation of the positive relationship between the tropical Atlantic sea
31 surface temperature (SST) and Korean precipitation during boreal summer season during 1900–
32 2010 is examined. The 15-yr moving correlation between the Tropical Atlantic SST (TAtlSST)
33 index (SST anomalies from 30° S to 30° N and 60° W to 20° E) and Korean precipitation
34 (precipitation anomalies from 35–40° N and 120–130° E) during June-July-August (JJA) exhibits
35 strong inter-decadal variation, which becomes positive at the 95% confidence level after the
36 1980s. Intensification of the linkage between the TAtlSST index and Korean precipitation after
37 the 1980s is attributed to global warming via the increased background SST. The increase in the
38 background SST over the Atlantic provides background conditions that enhance anomalous
39 convective activity by anomalous Atlantic SST warming. Therefore, the overall atmospheric
40 responses associated with the tropical Atlantic SST warming could intensify.

41 The correlation between the TAtlSST index and Korean precipitation also exhibits strong
42 inter-decadal variation within 1980–2010, which is over 0.8 during early 2000s, while it is
43 relative low (i.e., around 0.6) during the early 1980s. The enhanced co-variability between the
44 tropical and the mid-latitude Atlantic SST during the early 2000s induces a TAtlSST-related
45 Rossby wave source over the mid-latitude Atlantic, which excites stationary waves propagated
46 from the Atlantic to the Korean peninsula across northern Europe and northeast Asia. This
47 Rossby-wave train induces a cyclonic flow over the northern edge of the Korea, which
48 intensifies southwesterly and results in precipitation over Korea. This observed decadal
49 difference is well simulated by the stationary wave model experiments with a prescribed
50 TAtlSST-related Rossby wave source over the mid-latitude Atlantic.

51

52 **1. Introduction**

53 The major rainy season during boreal summer in Korea, which has two rainfall-peaks
54 (between late June and late July, and between mid-August and early September), is called
55 *Changma* (Wang et al., 2007; Ho and Kang, 1988; Lim et al., 2002; Seo et al., 2011; Lee et al.,
56 2017). As an important component of the East Asia summer monsoon (EASM) system, the
57 climatological evolution of Changma is closely linked to the summer monsoon rainfall band
58 across east Asia, which leads major rainfall in China (i.e., Mei-yu) and Japan (i.e., Baiu).
59 However, the variability of Changma is known to be rather different from Mei-yu or Baiu due to
60 its more complex mechanisms (Seo et al., 2011; Jeong et al., 2017).

61 While inter-annual variation of rainfall anomalies during boreal summer in China or Japan is
62 known to be determined by the large-scale sea surface temperature (SST) over the tropical
63 Pacific (e.g., the El Niño Southern Oscillation: ENSO; Huang and Wu, 1989; Wu et al., 2010), it
64 is not clear whether the Pacific SST can lead to variation in summer rainfall in Korea (Chen and
65 Zhou, 2014; He and Zhou, 2014). For example, Yun et al. (2014) used empirical orthogonal
66 function (EOF) analysis to show that the first EOF of the summer-averaged convection
67 anomalies over Asia is controlled by the tropical Pacific SST, and that the first EOF exhibits a
68 contrast pattern in the convection activity between China and Japan, while the convection
69 anomalies over Korea are relatively weaker. In addition, Son et al. (2014, 2015) found that a
70 significant statistical relationship between Korean precipitation and the ENSO is shown during
71 the September and boreal-winter development of El Nino, while the summer Korean
72 precipitation is weakly correlated to the ENSO.

73 The role of the Indian Ocean on the summer monsoon over East Asia is known to some
74 extent. Basin-wide Indian Ocean warming is reported to affect the inter-annual variability of the

75 rainfall over the western north Pacific and Japan through the Pacific-Japan (PJ) pattern (Kosaka
76 and Nakamura, 2006; Kosaka et al., 2013). This is further supported by the Atmospheric Global
77 Climate Model (AGCM) experiments (Annamalai et al., 2005; Yang et al., 2007). However, the
78 relationship between the PJ pattern and Korean precipitation is known to be marginal (Kosaka et
79 al., 2013). The intensity of Changma is controlled by the intensity of the Indian summer
80 monsoon (ISM) by energy variation of the circumglobal teleconnection (CGT) pattern (Ding et
81 al., 2011; Wang et al., 2012; Lee and Ha, 2015).

82 While the dynamic role of the Pacific or Indian Ocean on the summer precipitation
83 variability over Korea has been actively investigated for the past several decades, studies on the
84 role of the Atlantic Ocean only started in recent years. Recently, Ham et al. (2017) found a strong
85 relationship between Korean precipitation and the tropical Atlantic SST during boreal summer.
86 They showed that the relationship is quite robust from early to late summer. They argued that the
87 positive Atlantic SST anomaly could induce a reduction in the convective anomalies over the
88 western Pacific by modulating the Walker Circulation. This excites the low-level anti-cyclonic
89 flow over the western north Pacific (i.e., Western-North Pacific Subtropical High, WNPSH). This
90 leads the southerly to advect wet air to Korea, which enhances precipitation. This atmospheric
91 bridge is quite different from the atmospheric bridge from the north Atlantic to Eurasia, which
92 involves the Rossby wave train from the north Atlantic to Eurasia (Wu et al., 2010, 2011; Lim,
93 2015; Ye et al., 2015).

94 This relationship between the tropical Atlantic SST and Korean summer precipitation varies
95 strongly on a decadal time scale. In Lee et al. (2017), it is pointed out that the 15 yr moving
96 correlation between the tropical Atlantic SST and Korean summer precipitation was negative
97 during the early 1900s, but showed a sudden change to exhibit a significant positive relationship

98 after the 1980s (see their Fig. 9, and also Fig. 1a in this paper). Specifically, the correlation is >
99 0.8 for around the 2000s, which denotes that more than 60% of the total Korean-precipitation
100 variability might be explained by the TAtlSST variability during boreal summer. However, the
101 dynamic mechanism of this strong inter-decadal variation of the Tropical Atlantic-Korea (TA-K)
102 teleconnection pattern has never before been examined.

103 Therefore, this study investigates the decadal modulation of the tropical Atlantic-Korea
104 teleconnection pattern using reanalysis data. In section 2, observational data and the model used
105 in this study are summarized. Section 3 provides the mechanism of the decadal-variation of the
106 Tropical Atlantic-Korea (TA-K) teleconnection pattern. A summary and discussion are provided
107 in Section 4.

108

109 **2. Data and Models**

110 *a. Observational data*

111 In this study, we used two datasets for precipitation. For land precipitation, the data used for
112 the Korean Peninsula precipitation index (Figure 1) during 1901–2011 was from the Climate
113 Research Unit (CRU, Harris et al., 2013). The Global Precipitation Climatology Project (GPCP,
114 Adler et al., 2003) precipitation data during 1979–2008 was used for linear regression analysis.
115 The observed sea surface temperature (SST) data during 1901–2011 was from the NOAA
116 Extended Reconstructed Sea Surface Temperatures version 3 (ERSST V3b, Smith et al., 2008).
117 The 500 hPa vertical pressure velocity data, 850 hPa zonal and meridional wind data during
118 1948–2010, and 500 hPa geopotential height data during 1979–2008 were from the National
119 Center for Environmental Prediction/National Center for Atmospheric Research (NCEP/NCAR)

120 reanalysis version 1 (NCEP1, Kalnay et al., 1996). All data were linearly de-trended before
121 analysis.

122

123 *b. The Stationary Wave Model (SWM)*

124 We use the SWM to investigate large-scale atmospheric response to the Rossby wave
125 sources found in the North Atlantic. This SWM is the dry dynamical core of a fully nonlinear
126 time-dependent baroclinic model (Ting and Yu, 1998; Schubert et al., 2011). It has three-
127 dimensional (3-D) primitive equations in σ coordinates with 14 unevenly spaced vertical levels.
128 Horizontally, the model has rhomboidal wavenumber-30 truncation. A rigid-lid boundary
129 condition is applied at the top and at the surface of the model atmosphere. For damping,
130 Rayleigh friction and Newtonian cooling are applied in the vorticity, divergence, and temperature
131 equations to ensure meaningful solutions (Ting and Yu, 1998; Lim, 2015).

132 Vorticity induced by divergent (irrotational) flow forces the model to generate a large-scale
133 rotational component of the atmospheric circulation in the form of Rossby wave propagation. In
134 our study, we estimated the regressed Rossby wave source in the TAtlSST index using
135 NCEP/NCAR re-analysis data. The SWM is forced by this Rossby wave source. The vertical
136 profile of the Rossby wave source has a maximum in the upper-troposphere (Liu et al., 1998).
137 The background basic state for the model is given by three dimensions globally from surface to
138 tropopause. With these configurations, the model generates large-scale wave trains across
139 Eurasia in response to the Rossby wave source in the North Atlantic (e.g., Fig. 8). For further
140 details of the model, see Ting and Yu (1998).

141

142 **3. Mechanism of the inter-decal variation of the Tropical Atlantic-Korea (TA-K)**

143 **teleconnection pattern**

144 Figure 1a shows the 15-yr moving correlation between the precipitation anomaly over the
145 Korean peninsula from 35–40° N and 120–130° E, and the Tropical Atlantic SST (TAtlSST)
146 index, defined as an area-averaged sea surface temperature (SST) from 30° S to 30° N and 60°
147 W to 20° E, during Korean summer (June-July-August: JJA). During the first half of the 20th
148 century, the correlation between Korean precipitation and the TAtlSST index varies from –0.5 to
149 zero at the 95% confidence level. For example, the correlation for 1910–1920 is negative, while
150 the correlation for 1920–1950 is nearly zero. The correlation during the 1960s is positive, but
151 becomes nearly zero for 1970–1980. However, after the mid-1980s, the correlation changed
152 abruptly to a significantly positive value at > 95% confidence. This indicates that the positive
153 relationship between the tropical Atlantic SST and Korean precipitation during boreal summer
154 became significant after the 1980s. To demonstrate clearly the decadal variation of the
155 relationship between the TAtlSST and Korean precipitation, in Figure 1b (and with black dots in
156 Fig. 1c) the scatter plot shows Korean precipitation and the TAtlSST index for 1948–1978 and
157 1979–2010, respectively. During 1948–1978, the correlation between the TAtlSST index and the
158 Korean precipitation is only –0.19, while that during 1979–2010 is 0.58 with > 95% confidence.
159 Hereafter, the periods 1948–1978 and 1979–2010 are denoted P1 and P2.

160 It is also interesting that the maximum 15-yr moving correlation is slightly > 0.8 around
161 2000 (red dots in Fig. 1a). This means that about 65% of the total variability in precipitation over
162 Korea during boreal summer occurs during this period. The red and blue dots in Fig. 1c show
163 that the correlation between the TAtlSST index and Korean precipitation during 1994–2008 is >
164 0.8, while that during 1979–1993 is 0.63. This means that the relationship between the tropical
165 Atlantic and Korean precipitation also varies after 1980s even though it continuously exhibits

166 significantly positive values. To assess the details of mechanisms of the decadal variation of the
167 Tropical Atlantic-Korea (TA-K) teleconnection pattern after 1980s, the period 1979–1993 is
168 denoted P3, while 1994–2008 is denoted P4.

169

170 *a. Decadal difference between P1 and P2*

171 To demonstrate the systematic difference associated with the TAtlSST index between P1 and
172 P2, Figure 2 shows the regression of the SST, vertical pressure velocity at 500 hPa, wind-vector
173 and stream function at 850 hPa during 1948–1978 (P1) and 1979–2010 (P2). During P1, the
174 regressed SST anomalies show a positive signal with anomalous upward motion over the
175 equatorial Atlantic (Fig. 2a, 2b). The anomalous upward motion is closely linked to the release of
176 latent heat by the condensation of water vapor, which heats the atmospheric and induces Gill-
177 type atmospheric circulation anomalies (Gill, 1980). As a result, low-level cyclonic circulations
178 are induced north and south (a pair) of the equatorial Atlantic. The weaker cyclonic circulation
179 over the region south of the equatorial Atlantic is probably associated with the location of the
180 anomalous convection, which is north of the equator with the northward shift of the Atlantic
181 Intertropical Convergence Zone (ITCZ) during boreal summer.

182 Ham et al. (2017) showed that the SST warming over the tropical Atlantic triggers the low-
183 level easterly wind anomaly over the maritime continent by modulating the zonal Walker
184 circulation. This leads the low-level divergence over the western Pacific to decrease the
185 precipitation anomalies. The negative atmospheric heating anomaly excites a low-level anti-
186 cyclonic circulation over the sub-tropical western Pacific. This is consistent with many previous
187 studies indicating that the tropical Atlantic SST can lead the equatorial easterly over the western
188 Pacific (Ding et al., 2010; Rong et al., 2010; Ham et al., 2013a,b). Consistent with previous

189 studies, an anti-cyclonic circulation is shown over the western-north Pacific (Figure 2c).

190 During P2, the positive SST anomalies over the Atlantic extend farther to the north
191 compared to those during P1. For example, the positive SST anomalies during P1 are nearly zero
192 north of 15° N, while the positive SST anomalies related to the TAtlSST index during P2 extend
193 to 30° N (Fig. 2d). In addition, the positive SST anomalies are clear over the mid-latitude and
194 north Atlantic, with weak negative SST anomalies over the mid-latitude western Atlantic.
195 Consistent with the positive SST anomalies over the tropical Atlantic, local anomalous upward
196 motion is shown clearly during P2 (Fig. 2e). The dramatic differences from P1 during P2 are the
197 responses of the atmospheric circulation over the Pacific. During P2, the low-level anti-cyclonic
198 flow associated with the TAtlSST index is well-induced over the western Pacific (Fig. 2f). The
199 low-level anti-cyclonic circulation leads the southerly over the Korean peninsula; then, induces
200 the anomalous upward motion that is the signal of increased precipitation.

201 The enhancement of the anti-cyclonic flow over the western Pacific associated with the
202 tropical Atlantic SST during P2 is consistent with previous studies (Rodionov, 2004; Hong et al.,
203 2014). Hong et al. (2014) revealed that the tropical Atlantic SST contributes significantly to the
204 intensified variability of the western-north Pacific subtropical high (WNPSH) after the early
205 1980s. Hong et al. (2014) also argued that the weakened positive relationship between the ENSO
206 and the SST over the tropical Atlantic is responsible for the intensified role of tropical Atlantic
207 SST on the WNPSH after the early 1980s. That is, when there is strong coupling between El
208 Nino and the tropical Atlantic SST, the low-level anti-cyclonic flow over the subtropical western
209 Pacific (induced by the tropical Atlantic SST) is cancelled by the cyclonic flow due to the El
210 Nino signal over the equatorial eastern Pacific.

211 In addition, they suspected that the difference in changes of mean state between P1 and P2

212 might contribute to enhancing WNPSH with tropical Atlantic SST warming; however, they left
213 the resolution of this issue for future work. To examine this point in more detail, Figure 3 shows
214 the background state difference of the SST and vertical pressure velocity at 500 hPa between P2
215 and P1. Due to the global warming in recent decades, the overall background SST in P2 was
216 systematically higher than that in P1. In particular, the increase in the SST over the Atlantic,
217 Indian Ocean, and equatorial and subtropical western Pacific during P2 was more robust than in
218 other regions. This implies that the increase in the background SST associated with global
219 warming might be related to the enhanced TA-K teleconnection observed in recent decades. With
220 the aid of the increased SST during P2, the background upward motion was enhanced over the
221 equatorial Atlantic and western Pacific. On the other hand, the background upward motion over
222 the central Africa and the equatorial central Pacific, where the increase of the background SST is
223 relatively weak, was systematically weakened.

224 To compare the background state difference related to the enhancement of the TA-K
225 teleconnection and global warming in more detail, Figure 4a and b show the linear trend of the
226 background SST and the vertical pressure velocity from 1948 to 2010. As shown in Fig. 3, the
227 linear trend of increase in the background SST over the Atlantic, Indian Ocean, and western
228 Pacific is clear. In addition, the positive SST trend over the eastern Pacific, and the negative SST
229 trend over the north Pacific are also shown. The linear trend of the vertical pressure velocity
230 exhibits enhanced upward motion over the regions with greater positive trends in SST. For
231 example, the negative vertical pressure velocity trend (i.e., enhanced upward motion) is robust
232 over the Indian Ocean, western Pacific, and the Atlantic; while the positive vertical pressure
233 velocity trend is shown over the equatorial central Pacific, where the linear SST trend is
234 relatively weak.

235 Figure 4c and d show the regressed 15 yr moving background SST and the vertical pressure
236 velocity at 500 hPa over the 15 yr moving correlation between the TAtlSST index and Korean
237 precipitation. Therefore, the spatial distribution of the regression is understood to denote the
238 background state difference to amplify the correlation between the TAtlSST index and Korean
239 precipitation. Note that the regression was performed for the period 1948–2010 due to data
240 limitation. Also, the student t-test for the regression line is performed for Fig. 4c, and 4d, and the
241 area over 95% confidence level are only shaded. The positive signal in the background SST over
242 the Atlantic, Indian Ocean, and eastern Pacific is shown clearly, implying that increase in the
243 background SST over those regions can enhance the relationship between the tropical Atlantic
244 SST and Korean precipitation. In addition, the enhanced upward motion over the tropical
245 Atlantic, and Indian Ocean/maritime continent, and the weakened upward motion over the
246 equatorial central Pacific, is associated with enhanced TA-K teleconnection. Those general
247 features are quite similar to the spatial distribution of the linear trend. For example, the pattern
248 correlation between the spatial distribution of the linear trend and that related to the enhanced
249 TA-K teleconnection over the tropics (from 30° S to 30° N, and 0–360° E) is significant at > 99%
250 confidence (i.e., 0.64 and 0.80 for the SST and the vertical pressure velocity, respectively). This
251 implies that global warming possibly plays an important role in enhancing the linkage between
252 the tropical Atlantic and the climate variability over the Korean peninsula.

253 Given this, how can the change in background state due to global warming amplify the
254 atmospheric signals related to the tropical Atlantic SST? The impact of the background state on
255 the atmospheric variability over the tropics has been investigated in previous studies (Choi et al.,
256 2011; Chung and Li, 2013; Ham et al., 2013; Ham and Kug, 2012, 2015, 2016). Ham and Kug
257 (2015) showed that simulation of the ENSO-related precipitation among climate models is linked

258 closely to that in the tropical mean state. They argue that the warmer background SST over the
259 equatorial central-eastern Pacific enhances local anomalous convective activity during the El
260 Nino, implying a positive relationship between wetness in the background state and El Nino-
261 related precipitation anomalies. Consistent with the previous view, Choi et al. (2011) showed that
262 the decades with warmer background SST over the tropical western Pacific helped to induce
263 stronger local precipitation response to the SST anomaly; and therefore, led to much more
264 frequent central Pacific-type El Ninos. This implies that the warmer background state over the
265 tropical Atlantic can enhance anomalous convective activities and the resultant teleconnections
266 induced by the tropical Atlantic SST anomaly.

267 To examine whether the background SST can influence the upward motion response to the
268 SST anomaly, Fig. 5a shows a scatter diagram of the 15-yr moving SST and the anomalous
269 vertical pressure velocity regressed onto the SST anomaly over the tropical Atlantic. Note that
270 the regressed vertical pressure velocity anomaly indicates how strongly the atmospheric upward
271 motion is induced due to the given SST forcing (Choi et al., 2011; Ham and Kug, 2015). It is
272 clearly shown that the warmer background SST over the tropical Atlantic can lead stronger
273 anomalous upward motion (i.e., negative vertical pressure velocity) forced by unit SST anomaly.
274 With aids of higher background SST during P2, the overall atmospheric response to the TAtlSST
275 is systematically stronger than that during P1, confirming that the warmer background SST
276 during P2 provides a background state favorable to induction of a stronger local convection
277 response to the tropical Atlantic SST anomaly.

278 In addition to the background SST over the tropical Atlantic, it was found that the warmer
279 background SST over the warm pool region plays some role in amplifying the TA-K
280 teleconnection. Hong et al. (2013) clearly showed through idealized AGCM experiments that the

281 atmospheric response over the tropical Pacific to the tropical Atlantic SST anomaly is strongly
282 dependent on the location of the climatological warm pool. According to their experiments, the
283 positive precipitation response to the tropical Atlantic SST anomaly is robust where the idealized
284 warm pool is located. They argued that the climatological low-level convergence resulting from
285 the warm pool over the western Pacific could lead strong low-level convergence feedback;
286 therefore, the anomalous precipitation response could easily be amplified over the warm-pool
287 region (Wang, 2000). This indicates that the warmer climatology over the warm pool could also
288 amplify the Atlantic-induced teleconnection pattern over the Pacific.

289 To confirm this point, Figure 5b shows the scatter diagram between the 15 yr moving
290 background SST over the warm pool region (from 5° S to 10° N and 100–140° E) and the
291 anomalous zonal wind at 850 hPa over the western Pacific (from 5° S to 10° N and 130–160° E)
292 regressed onto the TAtlSST index. To some extent, that notion that the TAtlSST-related easterly
293 over the western Pacific tends to be enhanced during decades with warmer background state over
294 the warm-pool region is supported. The stronger easterly over the western Pacific is linked to the
295 anti-cyclonic flow over the subtropical western Pacific. This leads the southerly to increase the
296 precipitation over the Korean peninsula.

297 As a short summary, the background SST increase due to global warming can enhance the
298 TA-K teleconnection pattern by amplifying 1) the anomalous TAtlSST-related convective activity
299 over the tropical Atlantic, and 2) the TAtlSST-related wind response over the equatorial western
300 Pacific. The first feature is related to the increased background SST over the tropical Atlantic,
301 and the second feature is related to the increased SST over the warm-pool region.

302

303 *b. Decadal difference between P3 and P4*

304 Figure 6 shows the regressed JJA SST, precipitation, and 850 hPa wind-vector anomalies
305 onto the TAtlSST index during 1979–1993 (P3) and 1994–2008 (P4). The general SST warming
306 over the tropical Atlantic from 30° S to 30° N is clearly shown in both periods. The SST
307 warming over the equatorial eastern Atlantic from 5° S to 5° N was slightly stronger during P3.
308 On the other hand, the SST warming over the western Atlantic between 10–30° N was stronger
309 during P4, which lead stronger positive precipitation anomalies over the equatorial Atlantic. The
310 stronger TAtlSST-related precipitation anomalies over the equatorial Atlantic can lead to the
311 enhanced atmospheric responses over the western Pacific as discussed in a previous sub-section.
312 That is, the easterly anomalies over the equatorial western Pacific, along with the southerly along
313 the east coast of the China, are systematically stronger during P4. This could contribute to a
314 stronger positive precipitation anomaly over Korea during P4.

315 In addition, the tripolar SST pattern over the Atlantic (related to warming over the tropical,
316 north, and mid-latitude eastern Atlantic) and weak negative SST anomaly over the mid-latitude
317 western Atlantic are clear in both periods, implying strong co-variability between the tropical and
318 mid-latitude Atlantic SST (Lau and Nath, 2001; Czaja and Marshall, 2001; Czaja and
319 Frankignoul, 2002; Huang and Shukla, 2005; Li et al., 2007). The TAtlSST-related SST warming
320 over the mid-latitude eastern Atlantic was enhanced during P4 more than during P3, indicating
321 that the co-variability between the tropical and the mid-latitude Atlantic became stronger during
322 P4. The SST warming during P4 reached > 2 °C in the mid-latitude eastern Atlantic, while it was
323 between 1 and 2 °C during P3. As a result, the enhanced SST warming over the mid-latitude
324 Atlantic during P4 increased the local precipitation anomalies. That is, the precipitation
325 anomalies regressed onto the TAtlSST index appeared negative over the mid-latitude Atlantic
326 from 30–40° N during P3 (red box in Fig. 6b), while the positive precipitation anomalies became

327 more apparent during P4 (red box in Fig. 6d).

328 Those differences in the TAtlSST-related precipitation anomalies over the mid-latitude
329 Atlantic could modify the teleconnection originating from the high-latitude wave source over the
330 Atlantic. To examine the high-latitude teleconnection related to the TAtlSST index during P3 and
331 P4, the geopotential height and the wave activity flux at 500 hPa related to the TAtlSST index
332 was calculated for P3 and P4 (Fig. 7). The wave activity flux was calculated as follows (Takaya
333 and Nakamura, 2001; Kosaka and Nakamura, 2006).

$$334 W = \frac{p \cos\phi}{2|U|} \left(\begin{array}{l} \frac{U}{a^2 \cos^2 \phi} \left[\left(\frac{\partial \psi'}{\partial \lambda} \right)^2 - \psi' \frac{\partial^2 \psi'}{\partial \lambda^2} \right] + \frac{V}{a^2 \cos \phi} \left[\frac{\partial \psi'}{\partial \lambda} \frac{\partial \psi'}{\partial \phi} - \psi' \frac{\partial^2 \psi'}{\partial \lambda \partial \phi} \right] \\ \frac{U}{a^2 \cos^2 \phi} \left[\frac{\partial \psi'}{\partial \lambda} \frac{\partial \psi'}{\partial \phi} - \psi' \frac{\partial^2 \psi'}{\partial \lambda \partial \phi} \right] + \frac{V}{a^2} \left[\left(\frac{\partial \psi'}{\partial \phi} \right)^2 - \psi' \frac{\partial^2 \psi'}{\partial \phi^2} \right] \\ \frac{f_0^2}{N^2} \left\{ \frac{U}{a \cos \phi} \left[\frac{\partial \psi'}{\partial \lambda} \frac{\partial \psi'}{\partial z} - \psi' \frac{\partial^2 \psi'}{\partial \lambda \partial z} \right] + \frac{V}{a} \left[\frac{\partial \psi'}{\partial \phi} \frac{\partial \psi'}{\partial z} - \psi' \frac{\partial^2 \psi'}{\partial \phi \partial z} \right] \right\} \end{array} \right) + C_U M \quad (1)$$

335 , where U, V denote the 15-yr averaged zonal, meridional wind, respectively. And, ψ' , a , λ , ϕ , z
336 denote the streamfunction regressed onto TAtlSST index, a radius of the earth, longitude, latitude,
337 and vertical direction, respectively. The zonal, and the meridional components of the wave
338 activity flux denote the zonal, and the meridional advection of the wave-activity (angular)
339 pseudomomentum (Takaya and Nakamura, 2001).

340 During P3, the negative geopotential height anomaly over Europe and eastern Russia and the
341 positive geopotential height anomaly over western Russia, were clear. The geopotential height
342 anomalies over Korea exhibited slightly positive values. The wave activity flux indicates that
343 geopotential height anomalies over Europe and the Russia were excited by the wave source over
344 the mid-latitude Atlantic. For example, the wave activity flux is predominantly northeastward
345 from the mid-latitude Atlantic to eastern Russia. This shows that the TAtlSST-related mid-
346 latitude wave source over the Atlantic could have excited a Rossby wave train over Europe and
347 Russia during P3. However, the positive geopotential height anomalies over the Korean

348 peninsula are not likely to be related to the Rossby wave train from the Atlantic to Russia. For
349 example, the wave activity flux over Korea is mostly eastward, while the Rossby wave train
350 originating from the wave source over the mid-latitude Atlantic is located north of Korea.

351 The Rossby wave train excited by conditions in the mid-latitude Atlantic, is also clear during
352 P4. The wave activity flux is northeastward from the mid-latitude Atlantic to Scandinavia, and it
353 flows to eastern Russia and northeast Asia. The geopotential height anomalies exhibit negative
354 values over Scandinavia and northwest Asia, and positive values over the eastern Atlantic and
355 western Russia. The Korean peninsula is affected by the negative geopotential height anomaly
356 over northeast Asia; therefore, the geopotential anomaly induces the increased precipitation
357 anomalies. In addition, the negative geopotential height anomaly north of the Korean peninsula
358 leads increased precipitation by inducing a southwesterly wind over Korea. This implies that the
359 Rossby wave train from the mid-latitude Atlantic to the Korean peninsula becomes apparent
360 during P4, which helps to enhance the positive relationship between the TAtlSST and Korean
361 precipitation during boreal summer.

362 To clearly show that the wave source over the mid-latitude Atlantic is critical to the Rossby
363 wave train from the Atlantic to the Korean peninsula during P4, we performed model
364 experiments using SWM by prescribing the Rossby wave source during P3 and P4 (Schubert et
365 al., 2011). The linearized form of the Rossby wave source (RWS) was calculated as follows (Lim,
366 2015).

367
$$RWS_L = -V'_\chi \cdot \nabla(\bar{\zeta} + f) - (\bar{\zeta} + f) \nabla \cdot V'_\chi - \zeta' \nabla \cdot \bar{V}_\chi - \bar{V}_\chi \cdot \nabla \zeta' \quad (2)$$

368 where the overbar denotes the 15-yr averaged value and the prime denotes the anomaly
369 associated with the TAtlSST index. The first and fourth term on the right hand of the equation are
370 associated with vorticity advection, while the second and third terms involve the generation of

371 waves by wind convergence/divergence, and the vorticity stretching induced by baroclinic wind
372 anomalies, respectively. In this study, the linearized RWS associated with the
373 convergence/divergence forcing (i.e., second and third terms) is prescribed in the SWM model to
374 focus on the generation of the Rossby wave related to the TAtlSST-related precipitation
375 anomalies (Lim, 2015). In addition, to emphasize the role of the mid-latitude precipitation
376 anomalies, the TAtlSST-related convergence/divergence is prescribed only for 30-40°N and 45-
377 25°W.

378 Figure 8 shows the stream-function anomalies at 500 hPa simulated in SWM with the
379 prescribed mid-latitude RWS related to the TAtlSST index during P3 and P4. Note that only a
380 rotational component of the wind is provided as a result of the SWM, and a positive stream-
381 function (i.e., anti-cyclonic flow) is matched to the positive geopotential height due to the
382 geostrophic balance. With the prescribed forcing during P3, a stationary wave response only
383 shows over the North Atlantic and Western Europe, and the anti-cyclonic flow is weakly
384 simulated north of the Korean peninsula. The overall stream-function response during P3 is
385 much weaker than that during P4, which might be due to the fact that the RWS during P3
386 exhibits much localized pattern than that during P4, which eventually weakens the area-averaged
387 RWS amplitude within the forcing area (i.e. 30-40°N and 45-25°W) (not shown).

388 During P4, stationary wave responses are systematically stronger than during P3. The
389 propagation of the stationary waves from the mid-latitude Atlantic to the Korean peninsula is
390 clearly shown, along with the negative stream-function anomaly over Europe and positive
391 stream-function anomaly over western Russia. In addition, the negative stream-function anomaly
392 centered north of the Korean peninsula is also well-simulated, even though it is shifted slightly
393 north from the observed location. This means that the TAtlSST-related convergence/divergence

394 anomalies induced by the precipitation anomalies over the mid-latitude Atlantic during P4, can
395 act as a RWS to excite the stationary waves propagated from the mid-latitude Atlantic to the
396 Korean peninsula, which enhances the Tropical Atlantic - Korea (TA-K) teleconnection. The
397 additional SWM experiment by prescribing basic state during P3 and the RWS during P4
398 confirms this point that the difference in the RWS, rather than the difference in the basic state,
399 plays a critical role on enhancement of the stream-function response during P4.

400 Figure 9 shows a schematic diagram of the connection between the tropical Atlantic and
401 Korean precipitation during P3 and P4. There are two pathways (bridges) by which the tropical
402 Atlantic SST could affect Korean precipitation: tropical and extra-tropical. The tropical bridge
403 indicates that the tropical Atlantic warming affects Korean precipitation by inducing variability
404 over the tropical western Pacific. That is, tropical Atlantic warming leads the easterly over the
405 western Pacific by modulating zonal Walker circulation. East of the easterly center, a low-level
406 divergence is induced, which decreases the precipitation over the tropical western Pacific. This
407 excites the anti-cyclonic Rossby waves over the Philippine Sea as a Gill-type response, which
408 leads the southerly over Korea to increase precipitation. Between P3 and P4, the strength of this
409 tropical bridge is enhanced more during P4.

410 Second, the co-variability between the tropical and mid-latitude SST anomalies over the
411 Atlantic can cause the extra-tropical bridge between the TAtlSST and the Korean peninsula. The
412 TAtlSST-related SST anomalies over the Atlantic exhibit the tripolar SST pattern (Lau and Nath,
413 2001; Czaja and Marshall, 2001; Czaja and Frankignoul, 2002; Huang and Shukla, 2005; Li et al.,
414 2007), which indicate positive SST anomalies over the tropical, mid-latitude eastern, and
415 northern Atlantic. This co-variability between the tropical and mid-latitude, or north Atlantic
416 warming becomes stronger during P4 than during P3, which enhances positive precipitation

417 anomalies over the mid-latitude Atlantic during tropical Atlantic warming. The enhanced
418 precipitation anomalies acts as a Rossby wave source, which excites a stationary Rossby wave
419 train propagated from the Atlantic to the Korean peninsula. This extra-tropical wave train
420 generates cyclonic flow over the area north of Korea, which leads a southerly over Korea to
421 increase the precipitation.

422

423 **4. Summary and Discussion**

424 It was shown that the correlation between the TAtlSST index and Korean precipitation
425 exhibits strong inter-decadal variation, and becomes positive at $> 95\%$ confidence level after the
426 1980s. The intensification of the linkage between the TAtlSST index and Korean precipitation
427 after 1980s is attributed to global warming via the increased background SST. The linear trend of
428 the background SST exhibits stronger SST increase over the Atlantic, eastern Pacific, and the
429 warm-pool region, and it is found that the spatial distribution of SST increases can play a role in
430 enhancing the TA-K teleconnection. The increase in the background SST over the Atlantic
431 provides background conditions that enhance anomalous convective activity due to anomalous
432 SST warming (Choi et al., 2011; Ham and Kug, 2012, 2015, 2016). Therefore, the overall
433 atmospheric responses associated with tropical Atlantic warming are enhanced. In addition, the
434 background SST increases over the warm-pool region also can boost the TA-K teleconnection by
435 leading stronger, low-level convergence feedback (Wang et al., 2000; Hong et al., 2013), which
436 amplifies the TAtlSST-related atmospheric responses over the warm-pool regions.

437 The correlation between the TAtlSST index and Korean precipitation also exhibits strong
438 inter-decadal variation in 1980–2010, which reaches > 0.8 in 1994–2008, while it is relatively
439 low (i.e., about 0.6) during 1979–1993. It is found that the mid-latitude teleconnection pattern

440 between the tropical Atlantic and the Korean peninsula becomes robust during 1994–2008 and
441 enhances the TA-K teleconnection. While the mid-latitude SST and precipitation anomalies
442 related to the TAtlSST index are relatively weak for 1979–1993, the co-variability between the
443 tropical and the mid-latitude SST is enhanced systematically in 1994–2008. This generates mid-
444 latitude Rossby wave sources that excite a propagation of stationary waves from the Atlantic to
445 the Korean peninsula through northern Europe and northeast Asia. Due to the stationary waves
446 from the high-latitude route, there is cyclonic flow at the northern edge of the Korean peninsula,
447 which induces an additional southwesterly that increases Korean precipitation. This is supported
448 by a stationary wave model (SWM) experiment prescribing a TAtlSST-related Rossby wave
449 source over the mid-latitude Atlantic during P3 and P4. The results imply that the mid-latitude
450 Rossby wave source is responsible for the observed decadal difference in the TAtlSST-related
451 teleconnection between P3 and P4.

452 The one of main findings in this study, that the role of inter-annual Atlantic SST variability
453 can be intensified due to global warming, is worth investigating in more detail in future work.
454 While the strength of Atlantic-induced climate variability over other ocean basins could be
455 enhanced due to the global warming through the mechanism proposed in this study; it is also
456 possible that the warming over the Atlantic in the future could be weakened because the strength
457 of Atlantic Meridional Overturning Circulation (AMOC) is expected to weaken in the future
458 (Mikolajewicz et al., 2007; Srokosz et al., 2012; Cheng et al., 2013; Deplazes et al., 2013).
459 Therefore, it is worthwhile to examine whether the active role of the Atlantic climate variability
460 is enhanced as a result of a direct global warming signal, or weakened due to the reduced AMOC
461 strength in the climate model simulations prescribing future emission scenarios.

462

463 **References**

464 Adler RF, Huffman GF, Chang A, Ferraro R, Xie P, Janowiak J, Rudolf B, Schneider U, Curtis
465 S, Bolvin D, Gruber A, Susskind J, Arkin P, Nelkin E (2003) The version 2 global
466 precipitation climatology project (GPCP) monthly precipitation analysis (1979-present). *J
467 Hydrometeorol* 4:1147–1167

468 Annamalai H, Liu P, Xie S-P (2005) Southwest Indian Ocean SST variability: its local effect and
469 remote influence on Asian monsoons. *J Clim* 18:4150–4167

470 Chen, X., and T. Zhou, 2014: Relative role of tropical SST forcing in the 1990s periodicity
471 change of the Pacific-Japan pattern interannual variability, *J. Geophys. Res. Atmos.*, **119**,
472 13,043–13,066

473 Cheng W, Chiang JCH, Zhang D (2013) Atlantic meridional overturning circulation (AMOC) in
474 CMIP5 models: RCP and historical simulations. *J Clim* 26:7187–7197

475 Chikamoto Y, Timmermann A, Luo J-J, Mochizuki T, Kimoto M, Watanabe M, Ishii M, Xie S-P,
476 Jin F-F (2015) Skilful multi-year predictions of tropical trans-basin climate variability. *Nat
477 Commun* 6:6869. doi:10.1038/ncomms7869

478 Choi J, An S-I, Kug J-S, Yeh S-W (2011) The role of mean state on changes in El Niño's flavor.
479 *Clim Dyn* 37:1205–1215

480 Chung P-H, Li T (2013) Interdecadal relationship between the mean state and El Niño types. *J
481 Clim* 26:361–379. doi:10.1175/JCLID-12-00106.1

482 Czaja A, Frankignoul C (2002) Observed impact of Atlantic SST anomalies on the North
483 Atlantic Oscillation. *J Clim* 15:606–623

484 Czaja A, Marshall J (2001) Observations of atmosphere-ocean coupling in the North Atlantic. *Q
485 J R Meteorol Soc* 127:1893–1916

486 Deplazes G, Lückge A, Peterson LC, Timmermann A, Hamann Y, Ka Hughen, Röhl U, Laj C,
487 Ma Cane, Sigman DM, Haug GH (2013) Links between tropical rainfall and North Atlantic
488 climate during the last glacial period. *Nat Geosci* 6(3):213–217. doi:10.1038/ngeo1712

489 Ding QH, Wang B, Wallace M, Branstator G (2011) Tropical-extratropical teleconnections in
490 boreal summer: observed interannual variability. *J Clim* 24:1878–1896

491 Ding R, Ha K-J, Li J (2010) Interdecadal shifts in the relationship between the East Asian
492 summer monsoon and the tropical Indian Ocean. *Clim Dyn* 34:1059–1071

493 Gill AE (1980) Some simple solutions for heat induced tropical circulations. *Q J R Meteorol Soc*

494 106:447–462

495 Ham Y-G, Chikamoto Y, Kug J-S, Kimoto M, Mochizuki T (2017) Tropical Atlantic-Korea
496 teleconnection pattern during boreal summer season. *Clim Dyn.* doi:10.1007/s00382-016-
497 3474-z

498 Ham Y-G, Kug J-S (2012) How well do current climate models simulate two types of El Niño?
499 *Clim Dyn.* doi:10.1007/s00382-011-1157-3

500 Ham Y-G, Kug J-S (2015) Role of north tropical Atlantic SST on the ENSO simulated using
501 CMIP3 and CMIP5 models. *Clim Dyn.* doi:10.1007/s00382-015-2527-z

502 Ham Y-G, Kug J-S (2016) ENSO amplitude changes due to greenhouse warming in CMIP5:
503 Role of mean tropical precipitation in the twentieth century. *Geophys Res Lett* 43.
504 doi:10.1002/2015gl066864

505 Ham Y-G, Kug J-S, Kim D, Kim Y-H, Kim D-H (2012) What controls phase locking of ENSO
506 to boreal winter in coupled GCMs? *Clim Dyn.* doi:10.1007/s00382-012-1420-2

507 Ham Y-G, Kug J-S, Park J-Y (2013) Two distinct roles of Atlantic SSTs in ENSO variability:
508 north Tropical Atlantic SST and Atlantic Niño. *Geophys Res Lett* 40:4012–4017.
509 doi:10.1002/grl.50729

510 Ham Y-G, Kug J-S, Park J-Y, Jin F-F (2013) Sea surface temperature in the north tropical
511 Atlantic as a trigger for El Niño/Southern oscillation events. *Nat Geosci* 6:112–116.
512 doi:10.1038/ngeo1686

513 Harris I, Jones PD, Osborn TJ, Lister DH (2013) Updated highresolution grids of monthly
514 climatic observations—the CRU TS3. 10 Dataset. *Int J Climatol.* doi:10.1002/joc.3711

515 He C, Zhou T (2014) The two interannual variability modes of the Western North Pacific
516 Subtropical High simulated by 28 CMIP5-AMIP models. *Clim Dyn* 43:2455–2469.
517 doi:10.1007/s00382-014-2068-x

518 Ho CH, Kang IS (1988) The variability of precipitation in Korea. *J Korean Meteorol Soc* 24: 38–
519 48

520 Hong CC, Chang TC, Hsu HH (2014) Enhanced relationship between the tropical Atlantic SST
521 and the summertime western North Pacific subtropical high after the early 1980s. *J Geophys*
522 *Res.* doi:10.1002/2013JD021394

523 Hong S, Kang I-S, Choi I, Ham Y-G (2013) Climate responses in the tropical Pacific associated
524 with Atlantic warming in recent decades. *Asia Pac J Atmos Sci* 49:209–217

525 Huang BH, Shukla J (2005) Ocean-atmosphere interactions in the tropical and subtropical
526 Atlantic Ocean. *J Clim* 18:1652–1672

527 Huang R, Wu Y (1989) The influence of ENSO on the summer climate change in China and its
528 mechanisms. *Adv Atmos Sci* 6:21–32

529 Jeong J-H, Lee H, Yoo JH, Kwon M, Yeh S-W, Kug J-S, Lee J-Y, Kim B-M, Son S-W, Min S-K,
530 Lee H, Lee W-S, Yoon J-H, Kim H-K (2017) The status and prospect of seasonal climate
531 prediction of climate over Korea and East Asia: A review. *Asia Pac J Atmos Sci* 53:149–173

532 Kalnay E, Kanamitsu M, Kistler R, Collins W, Deaven D, Gandin L, Iredell M, Saha S,
533 Woollen J, Zhu Y, Chelliah M, Ebisuzaki W, Higgins W, Janowiak J, Mo KC, Ropelewski
534 C, Wang J, Leetma A, Reynolds R, Jenne R, Joseph D (1996) The NCEP/NCAR 40 year
535 reanalysis project. *Bull Am Meteorol Soc* 77:437–471

536 Kosaka Y, Nakamura H (2006) Structure and dynamics of the summertime Pacific-Japan
537 teleconnection pattern. *Q J R Meteorol Soc* 132(619):2009–2030. doi:10.1256/qj.05.204

538 Kosaka Y, Xie S-P, Lau N-C, Vecchi GA (2013) Origin of seasonal predictability for summer
539 climate over the Northwestern Pacific. *Proc Natl Acad Sci.* doi:10.1073/pnas.1215582110

540 Kucharski F et al (2016) Atlantic forcing of Pacific decadal variability. *Clim Dyn.*
541 doi:10.1007/s00382-015-2705-z

542 Lau N-C, Nath MJ (2001) Impact of ENSO on SST variability in the North Pacific and North
543 Atlantic: seasonal dependence and the role of extratropical sea-air coupling. *J Clim* 14:
544 2846–2866

545 Lee J-Y, Ha K-J (2015) Understanding of Interdecadal Changes in Variability and Predictability
546 of the Northern Hemisphere Summer Tropical-Extratropical Teleconnection. *J Clim*
547 28:8634–8647. doi:10.1175/JCLI-D-15-0154.1

548 Lee J-Y, et al., (2017) The long-term variability of Changma in the East Asian summer monsoon
549 system: A review and revisit, Submitted to *Asia. Pac. J. Atmos.*

550 Li S, Robinson WA, Hoerling MP, Weickmann KM (2007) Dynamics of the extratropical
551 response to a tropical Atlantic SST anomaly. *J Clim* 20(3):560–574

552 Li X, Xie S-P, Gille ST, Yoo C (2016) Atlantic-induced pan-tropical climate change over the
553 past three decades. *Nat Clim Change* 6:275–279

554 Lim, Y.-K., K.-Y. Kim, and H.-S. Lee (2002) Temporal and spatial evolution of the Asian
555 summer monsoon in the seasonal cycle of synoptic fields. *J. Climate*, **15**, 3630–3644.

556 Lim Y-K (2015) The East Atlantic/West Russia (EA/WR) teleconnection in the North Atlantic:
557 climate impact and relation to Rossby wave propagation. *Clim Dyn* 44:3211–3222.
558 doi:10.1007/s00382-014-2381-4

559 Liu AZ, Ting M, Wang H (1998) Maintenance of circulation anomalies during the 1988 drought
560 and 1993 floods over the United States. *J Atmos Sci* 55:2810–2832

561 McGregor S et al (2014) Recent Walker circulation strengthening and Pacific cooling amplified
562 by Atlantic warming. *Nat Clim Change* 4(10):888–892

563 Mikolajewicz U, Vizcaino M, Jungclaus J, Schurgers G (2007) Effect of ice sheet interactions in
564 anthropogenic climate change simulations. *Geophys Res Lett* 34:L18706.
565 doi:10.1029/2007GL031173

566 Rodionov SN (2004) A sequential algorithm for testing climate regime shifts. *Geophys Res Lett*
567 31. doi:10.1029/2004GL019448

568 Rong X, Zhang R, Li T (2010) Impacts of Atlantic sea surface temperature anomalies on Indo-
569 East Asian summer monsoon-ENSO relationship. *Chin Sci Bull* 55:2458–2468

570 Schubert S, Wang H, Suarez M (2011) Warm season subseasonal variability and climate
571 extremes in the Northern Hemisphere: the role of stationary Rossby waves. *J Clim* 24:4773–
572 4792

573 Seo K-H, Son J-H, Lee J-Y (2011) A new look at changma: atmosphere. *Korean Meteor Soc*
574 21:109–121

575 Smith T, Reynolds R, Peterson T, Lawrimore J (2008) Improvements to NOAA's historical
576 merged land–ocean surface temperature analysis (1880–2006). *J Clim* 21:2283–2296.
577 doi:[10.1175/2007JCLI2100.1](https://doi.org/10.1175/2007JCLI2100.1)

578 Son H-Y, Park J-Y, Kug J-S, Yoo J, Kim C-H (2014) Winter precipitation variability over
579 Korean Peninsula associated with ENSO. *Clim Dyn* 42:3171–3186

580 Son HY, Park JY, Kug JS (2015) Precipitation variability in September over the Korean
581 Peninsula during ENSO developing phase. *Clim Dyn.* doi:10.1007/s00382-015-2776-x

582 Srokosz M, Baringer M, Bryden H, Cunningham S, Delworth T, Lozier S, Marotzke J, Sutton R
583 (2012) Past, present and future change in the Atlantic meridional overturning circulation.
584 *Bull Amer Meteorol Soc.* doi:10.1175/BAMS-D-11-00151.1

585 Steinman BA, Mann ME, Miller SK (2015) Atlantic and Pacific multidecadal oscillations and
586 Northern Hemisphere temperatures. *Science* 347(6225):988–991.

587 doi:10.1126/science.1257856

588 Takaya K, Nakamura H (2001) A formulation of a phase-independent wave-activity flux for
589 stationary and migratory quasigeostrophic eddies on a zonally varying basic flow. *J Atmos*
590 *Sci* 58:608–627

591 Wang B, Jhun J-G, Moon B-K (2007) Variability and singularity of Seoul, South Korea, rainy
592 season (1778–2004). *J Clim* 20:2572–2580

593 Wang B, Wu R-G, Fu X-H (2000) Pacific-east Asian teleconnection: How does ENSO affect
594 Asian climate? *J Clim* 13:1517–1536

595 Wang C (2000) On the atmospheric responses to tropical Pacific heating during the mature phase
596 of El Nino. *J Atmos Sci* 57:3767–3781

597 Wang H, Wang B, Huang F, Ding Q, Lee JY (2012) Interdecadal change of the boreal summer
598 circumglobal teleconnection (1958–2010). *Geophys Res Lett* 39:L12704.

599 Wu R, Yang S, Liu S, Sun L, Lian Y, Gao Z (2011) Northeast China summer temperature and
600 North Atlantic SST. *J Geophys Res* 116:D16116. doi:10.1029/2011JD015779

601 Wu R, Yang S, Liu S, Sun L, Yi L, Gao Z (2010) Changes in the relationship between Northeast
602 China summer temperature and ENSO. *J Geophys Res* 115:D21107.
603 doi:10.1029/2010JD014422

604 Yang J, Liu Q, Xie S-P, Liu Z, Wu L (2007) Impact of the Indian Ocean SST basin mode on the
605 Asian summer monsoon. *Geophys Res Lett* 34:L02708. doi:10.1029/2006GL028571

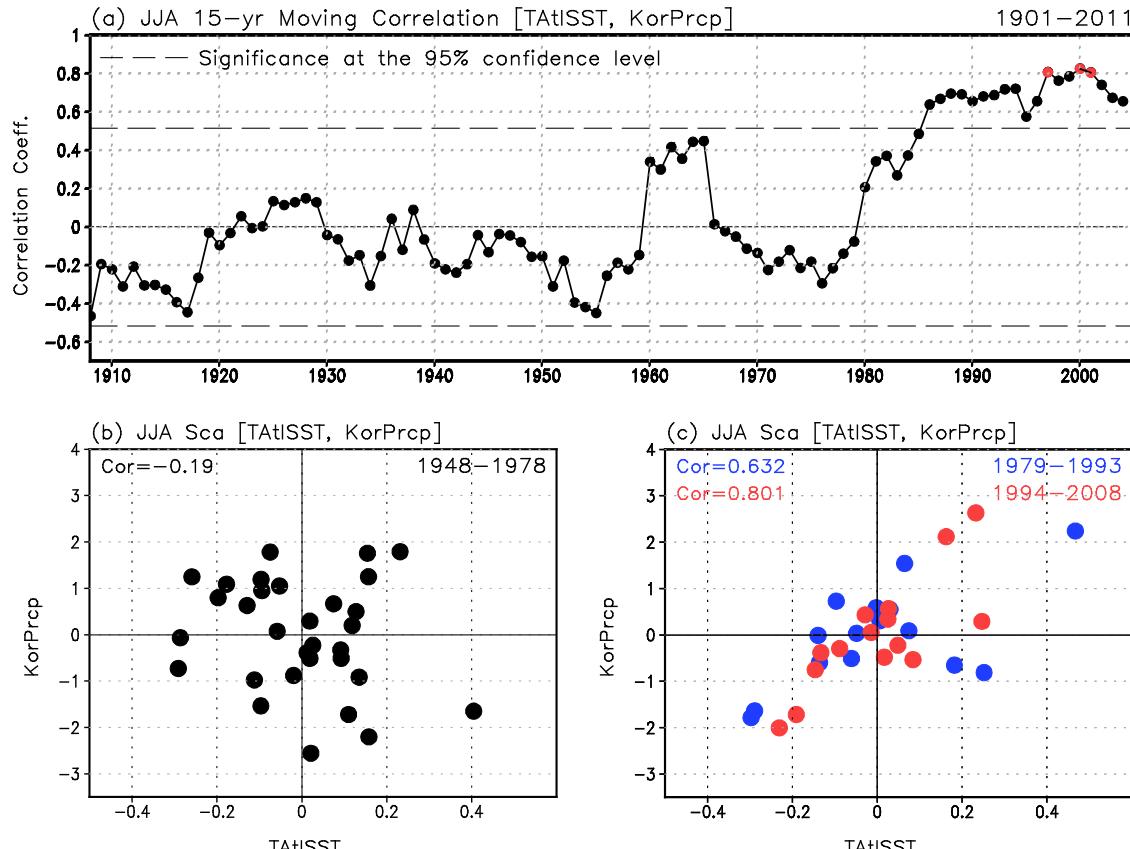
606 Ye K, Wu R, Liu Y (2015) Interdecadal change of Eurasian snow, surface temperature, and
607 atmospheric circulation in the late 1980s. *J Geophys Res* 120:2738–2753.
608 doi:10.1002/2015JD023148

609 Yun K-S, Lee J-Y, Ha K-J (2014) Recent intensification of the South and East Asian monsoon
610 contrast associated with an increase in the zonal tropical SST gradient. *J Geophys Res*
611 *Atmos* 119:8104–8116

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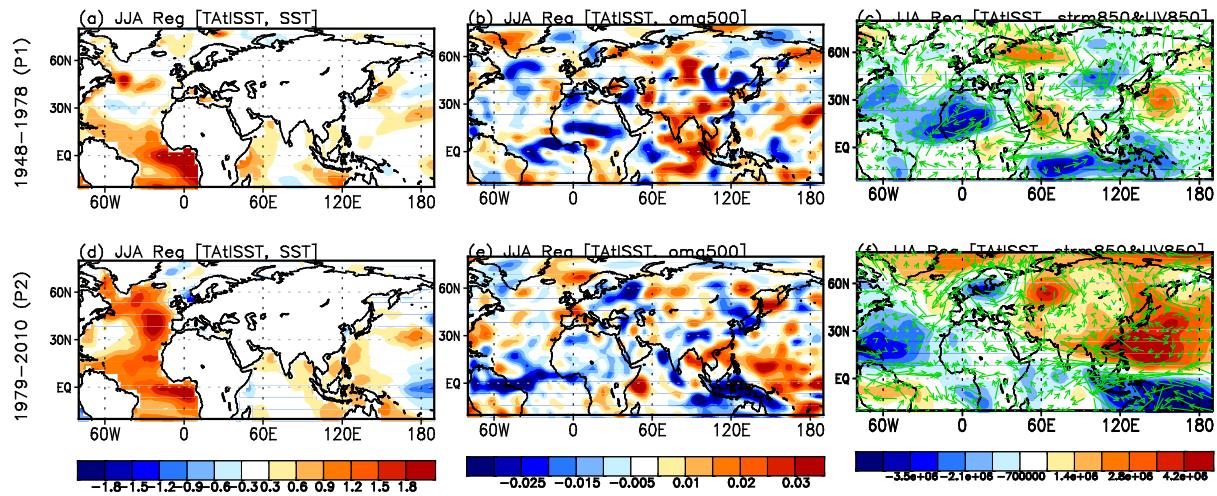
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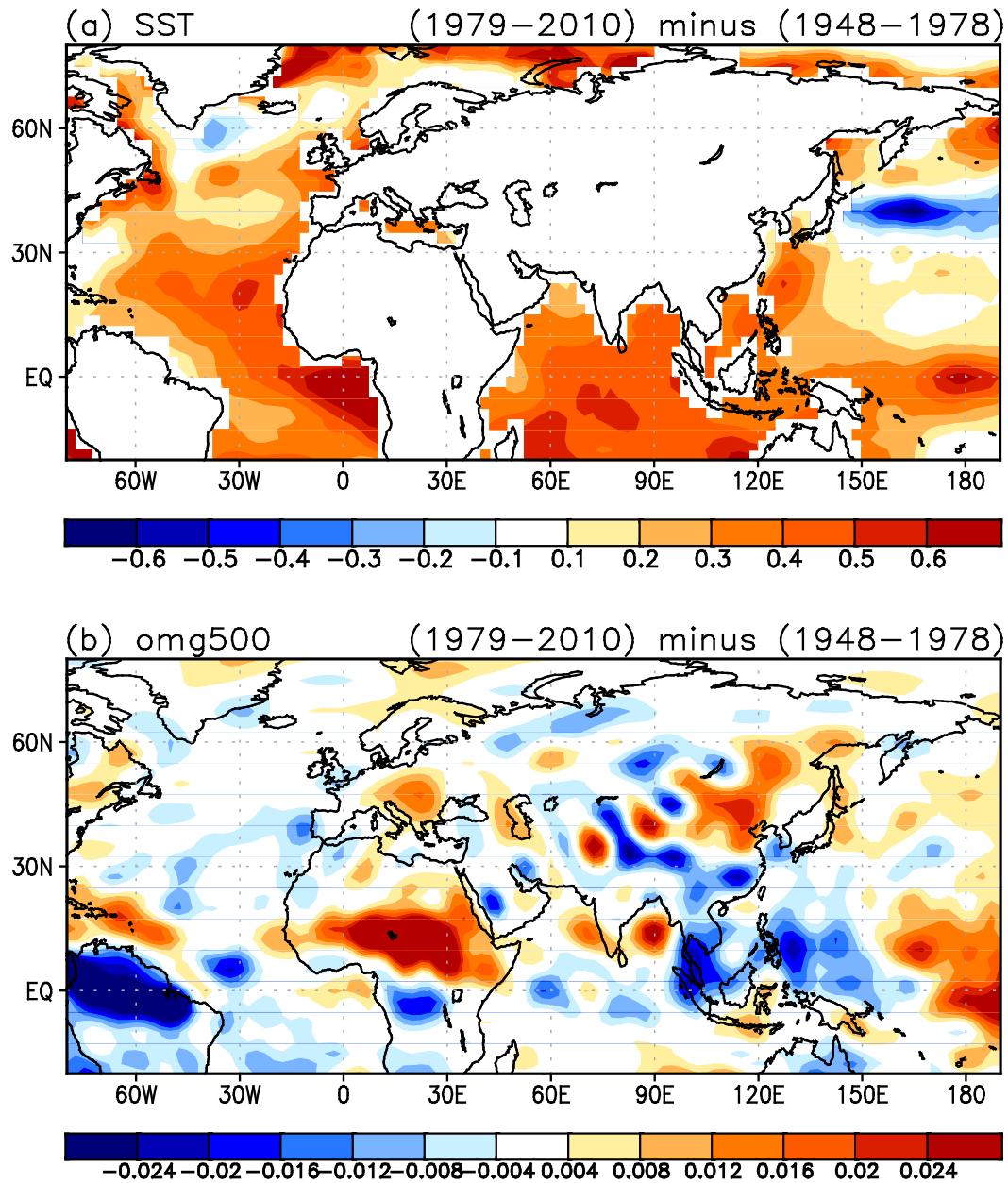
617 Figure 1. (a) The 15 yr moving correlation between precipitation anomalies over the Korean
 618 peninsula (35–40° N, 120–130° E) and Tropical Atlantic SST (60° W to 20° E and 30° S to 30°
 619 N; TAtISST) during Korean summer (June-July-August: JJA) from 1901 to 2011. The dashed
 620 line indicates statistical significance at the 95% confidence level by the Students' t-test. (b) A
 621 scatter diagram of the precipitation anomalies over the Korean peninsula and TAtISST anomalies
 622 during JJA, from 1948 to 1978. (c) As in (b), but the black dots indicate 1979–2010, the blue
 623 dots 1979–1993, and the red dots 1994–2008.

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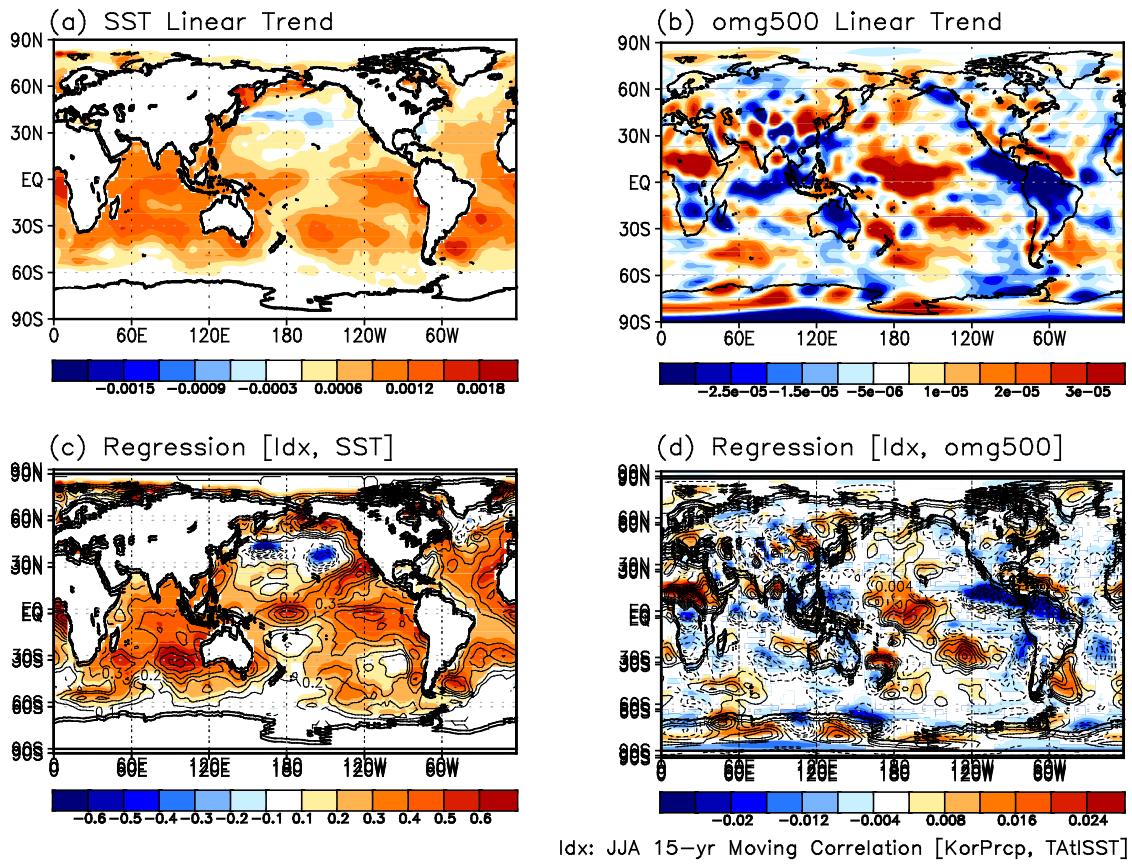
626
 627 Figure 2. Regression maps of (a, d) SST (K/K), (b, e) Vertical pressure velocity at 500 hPa
 628 (hPa/K), (c, f) Stream function (shaded, m²/s/K) and horizontal wind anomalies at 850 hPa
 629 (vector, m/s/K) onto the TAtlSST index during JJA of 1948–1978 (upper, P1) and 1979–2010
 630 (lower, P2).



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633 Figure 3. Difference in background (a) SST and (b) Vertical pressure velocity at 500 hPa during
634 1979–2010 (P2) from that during 1948–1978 (P1).

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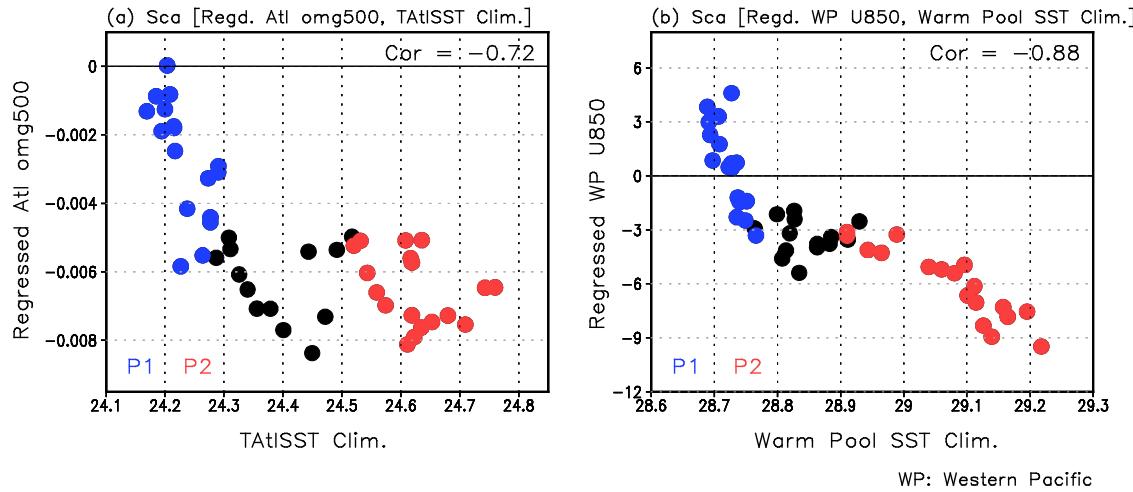


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638 Figure 4. Linear trend of the background (a) SST and (b) Vertical pressure velocity at 500 hPa
639 from 1948 to 2010: Linear regression maps of (c) SST and (c) Vertical pressure velocity at 500
640 hPa onto the 15-yr moving correlation between the TAtlSST index and Korean precipitation
641 from 1948 to 2010. In panel (c) and (d), the regions over 95% confidence level with student t-test
642 are shaded.

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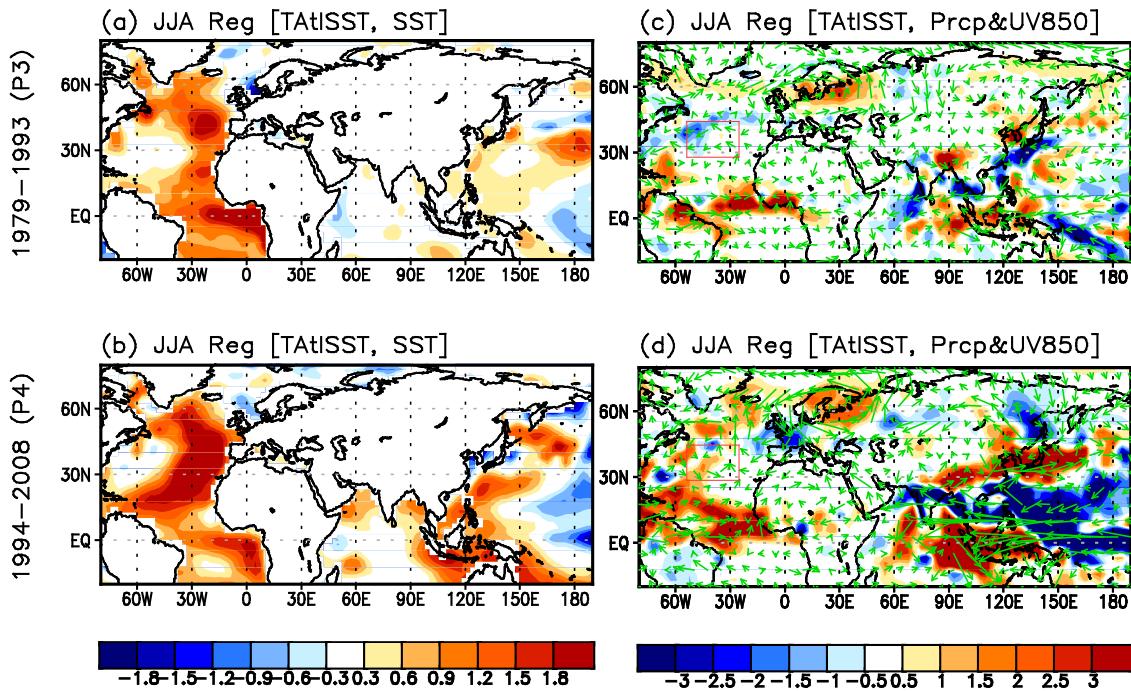
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 647 Figure 5. (a) A Scatter diagram between the 15-yr moving SST over the tropical Atlantic region
 648 (30° S to 30° N and 60° W to 20° E,) and the regressed vertical pressure velocity over the
 649 tropical Atlantic (20° S to 30° N and 60° W to 20° E) onto the TAtlSST index from 1948 to 2010.
 650 (b) A scatter diagram between the 15-yr moving SST over the warm pool region (5° S to 10° N
 651 and $100\text{--}140^{\circ}$ E,) and the regressed zonal wind at 850 hPa over the western Pacific (5° S to 10°
 652 N and $130\text{--}160^{\circ}$ E; WP) onto the TAtlSST index from 1948 to 2010.

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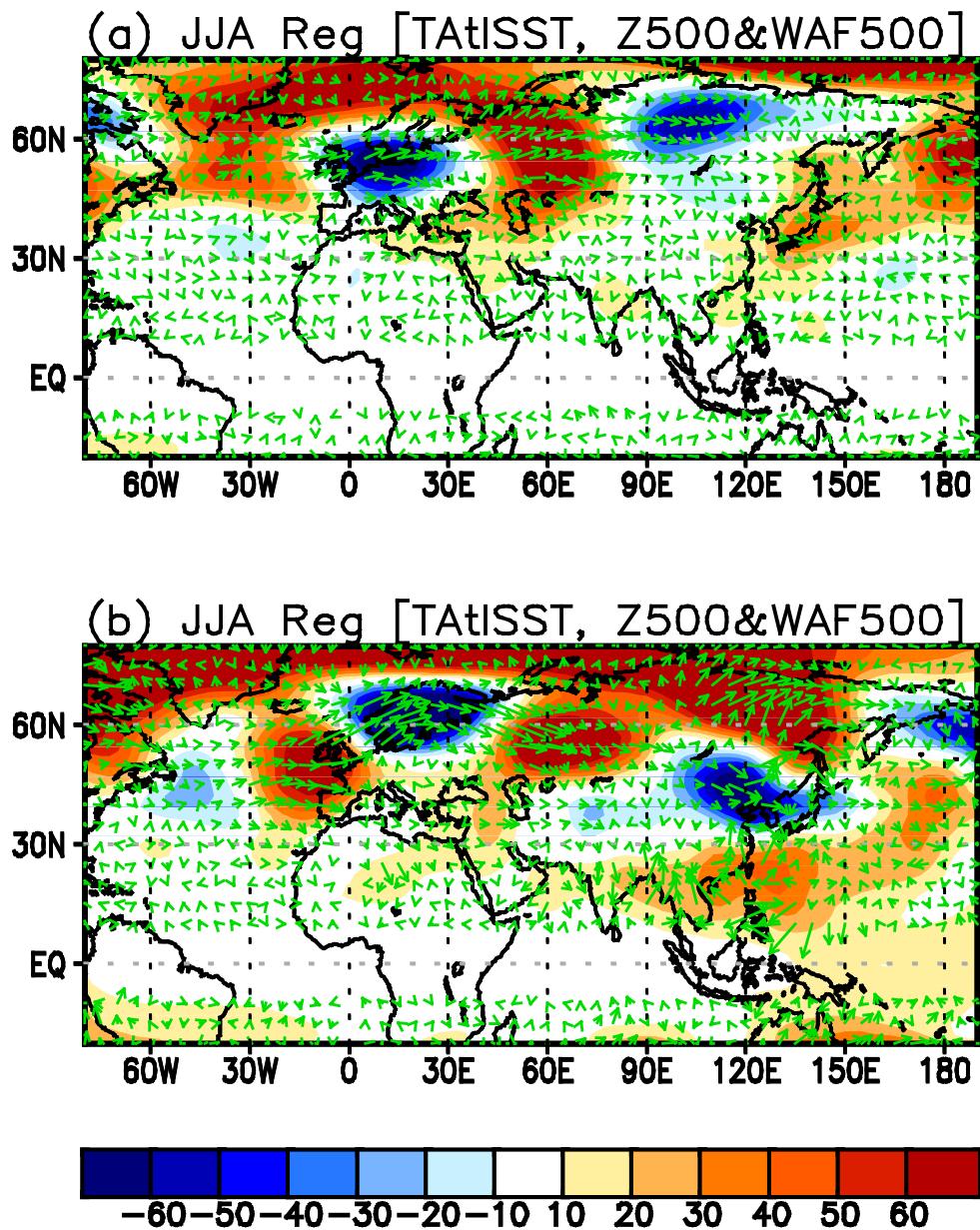


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 656 Figure 6. Regression maps of (a, b) SST (K/K); (c, d) Precipitation (shaded, mm/day/K) and
 657 horizontal wind anomalies at 850 hPa (vector, m/s/K) onto the TAtlSST index during JJA of
 658 1979–1993 (upper, P3) and 1994–2008 (lower, P4).

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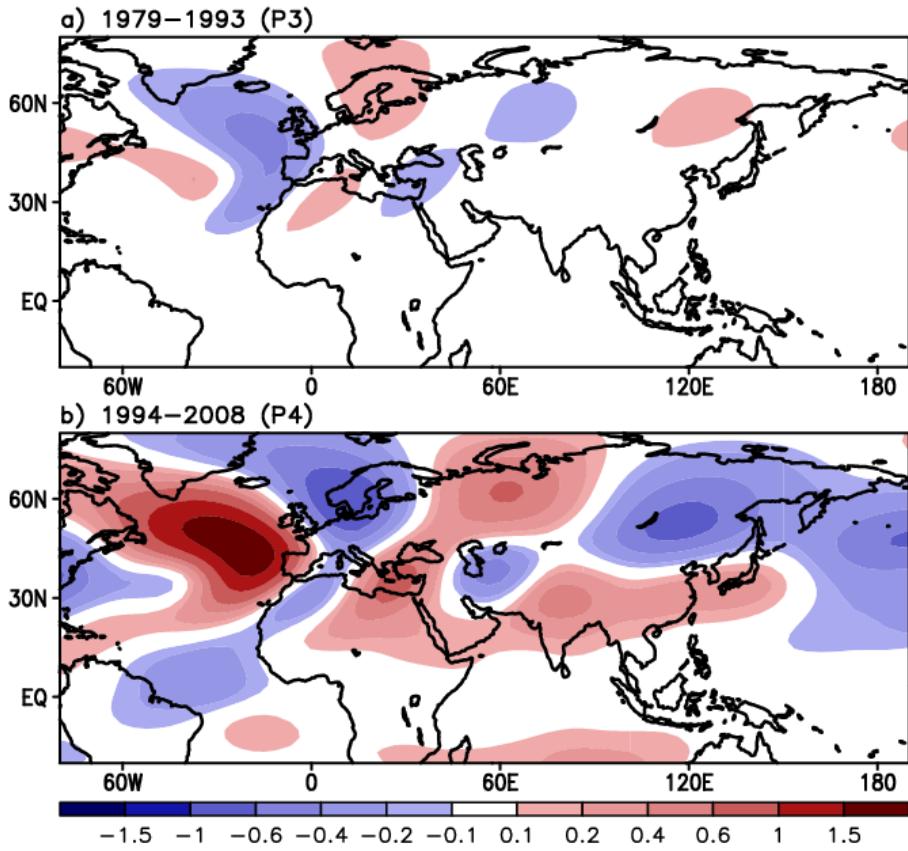
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 663 Figure 7. Regression maps of geopotential height at 500 hPa (shaded) and wave activity flux at
 664 500 hPa (vector) onto the TAtISST index during JJA of (a) 1979–1993 (P3) and (b) 1994–2008
 665 (P4).

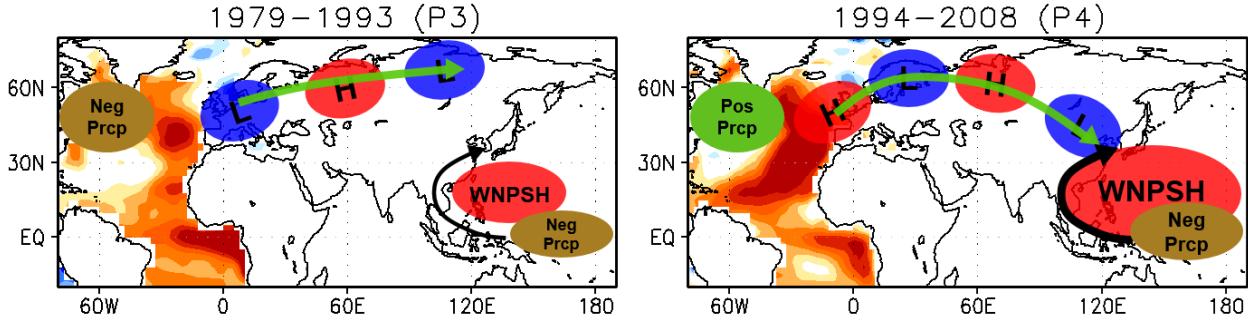
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668 Figure 8. Stream-function anomalies at 500 hPa simulated in a stationary wave model (SWM)
 669 with the prescribed mid-latitude Rossby wave source related to TAtlSST index during (a) 1979–
 670 1993 (P3) and (b) 1994–2008 (P4). The linearized RWS is prescribed only over the mid-latitude
 671 eastern Atlantic (30–40° N and 45–25° W).

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673
 674 Figure 9. Schematic diagram of the connection between the tropical Atlantic and Korean
 675 precipitation for 1979–1993 (P3) and 1994–2008 (P4). Shading over the Atlantic shows
 676 anomalous SST warming related to the TAtlSST index. The small red/blue circles show
 677 anomalous high/low pressure at 500 hPa, and the large red circle over the western Pacific
 678 represents the western-north Pacific subtropical high (i.e., WNPSH). The green/brown circles
 679 show anomalous positive/negative precipitation. The green vector represents the wave activity
 680 flux and black vector represents the low-level wind. The Rossby wave train from the mid-
 681 latitude Atlantic to Korean peninsula (green vector) becomes apparent along with the stronger
 682 intensity of the WNPSH during P4. Both lead a stronger southerly over the Korean peninsula
 683 (black vector) to enhance the TAtlSST-related precipitation anomalies over Korea during P4.
 684